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FOR THE COMMANDER

John Howard

Chief Scientist

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The use of a one-dimensional model atmosphere calculation incorporating dissociation of O_2 and ionized species demonstrates that dissociation, through the production of atomic oxygen, can drive the vertical thermosphere and upper mesosphere into diurnal oscillations. It is further demonstrated that neutral mesospheric and lower thermospheric turbulence, through its control of atomic oxygen flow, can cause large amplitude variations of these oscillations.

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Preface

The use of a one-dimensional model atmosphere calculation incorporating dissociation of O_2 and ionized species demonstrates that dissociation, through the production of atomic oxygen, can drive the vertical thermosphere and upper mesosphere into diurnal oscillations. It is further demonstrated that neutral mesospheric and lower thermospheric turbulence, through its control of atomic oxygen flow, can cause large amplitude variations of these oscillations.

Contents

Illustrations

Dissociation Driven Diurnal Oscillations

1. INTRODUCTION

The dynamic processes taking place among the neutral species in the thermosphere are of an extremely complex nature. The interplay of driving forces and atmospheric response has been considered in many theories and texts (to cite a few)[1, 2, 3] and many more in the recent literature. Generally, heat energy is considered as the driving mechanism of tides and motions in these theories. In the thermosphere, the EUV solar radiation below 1000 Å is absorbed above 120 km and heats the neutral gas through ionization, dissociation, and recombination of the plasma. Radiation at longer wavelengths is absorbed below 120 km and again there is the generation of heat energy through absorption, dissociation, and subsequent recombination of ground state and excited species[4, 5] that drives the atmosphere to

(Received for publication 30 May 1978)

1. Chapman, S., and Cowling, T.G. (1952) The Mathematical Theory of Non-Uniform Gases, 2nd ed. Cambridge University Press, New York.
2. Wilkes, M.V. (1949) Oscillations of the Earth's Atmosphere, Cambridge University Press, New York.
3. Harris, I., and Priester, W. (1962) Time-dependent structure of the upper atmosphere, J. Atmos. Sci. 19:286-301.
4. Lindzen, R.W., and Blake, D. (1970) Internal Gravity waves in atmosphere with realistic dissipation and temperature, 2, Geophys. Fluid. Dyn. 2:31
5. Volland, M., and Mayr, H.G. (1973) A numerical study of three-dimensional variations in the thermosphere, Ann. Geophys. 29:1102.

respond to this diurnal stimulus. We would like to discuss another mechanism that can induce atmospheric motions. It is ignored in most calculations but when incorporated it is generally not considered separately from the heat source. It is usually taken into account in one-dimensional models that try to display minor species and temperature distributions and diffusive motions in the lower thermosphere and ionosphere. This mechanism is the dissociation of molecular oxygen to produce oxygen atoms that through their creation and subsequent diffusive migration from their source will force the atmosphere into motion. This effect, which in our model exhibits a diurnal response to the diurnal driving force, is created when there is complete coupling among all species through collisions. Interactions, primarily atomic oxygen with N_2, start the molecular nitrogen into a diurnal up and down motion. This action continues until all species, neutral and ionic, react to this mean mass motion through collisions or through chemical interactions.

The theory we apply follows from Chapman and Cowling, (Reference 1) and Hirschfelder et al[6] for the equation of motion in a coupled multicomponent gas (also see Banks and Kockarts[7]). Expressed as one-dimensional equations in the vertical, there are the neutral motions

$$\frac{\partial C_i}{\partial t} + \frac{kT}{m_i}\left[\frac{1}{n_i}\frac{\partial n_i}{\partial z} + \frac{1}{T}\frac{\partial T}{\partial z} + \frac{1}{H_i}\right] = -\sum_j \nu_{ij}\,(C_i - C_j) \tag{1}$$

and the ion motions

$$\frac{\partial C_i}{\partial t} + \frac{kT}{m_i}\cdot\left[\frac{1}{n_i}\frac{\partial n_i}{\partial z} + \frac{1}{n_e}\frac{\partial n_e}{\partial z} + \frac{2}{T}\frac{\partial T}{\partial z} + \frac{1}{H_i}\right]$$

$$= -C_i\,\nu_i \left\{\frac{\Omega^2 + \nu_i^2}{\Omega^2 \sin^2 I + \nu_i^2}\right\} + \sum_j \nu_{ij}\,C_j\,. \tag{2}$$

In these equations

C_i is the velocity of the i th species,

n_i is the concentration of the i th species,

n_e is the electron concentration,

T is temperature,

k is Boltzmann's constant,

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6. Hirschfelder, J.O., Curtiss, C.F., and Bird, R.B. (1954) Molecular Theory of Gases and Liquids, John Wiley, New York.
7. Banks, P.M., and Kockarts, G. (1973) Aeronomy, Academic Press, New York.

ν_{ij} is the collision frequency of species i in species j, $\nu_i = \sum_j \nu_{ij}$,

Ω is the gyrofrequency of the electrons,

I is the magnetic dip angle.

These equations are coupled to the equation of continuity

$$\frac{\partial n_i}{\partial t} = F_i - n_i R_i - \frac{\partial \phi_i}{\partial z} \qquad (3)$$

and the mean mass motion (C_o) is determined from $C_o = \frac{\sum_i \rho_i C_i}{\sum_i \rho_i}$

where

F_i is the rate at which species i is chemically formed,

$n_i R_i$ is the rate at which species i is chemically removed,

$\phi_i = n_i C_i$ *is the flux of species i,*

ρ_i is the mass density of species i.

The initial conditions used in these calculations are the results discussed in Philbrick et al[8] that displays a one-dimensional model of the atmosphere using the temperature and mass density data as well as the turbulent diffusion coefficients (K and 3 K) determined from the ALADDIN I experiments. These Philbrick et al[8] data were then extrapolated from 150 km to 400 km using the diffusion equilibrium hypothesis. Further results giving the minutiae of the expanded ion-neutral model synopsized here will be presented in a more complete dissertation later. For this report, we shall present only some of the salient points supporting our thesis from the fifteenth model day when most of the major species have arrived at near diurnal reproducibility. These calculations cover the altitude region 50 to 400 km.

2. RESULTS AND CONCLUSIONS

The results of the calculation that display the diurnal forcing by the oxygen flow, that in turn drives the other constituents are presented in Figures 1, 2, 3, and 4. These show the diurnal variation of O, Ar, N_2, and ρ at specific altitudes in the mesosphere and thermosphere for the turbulent diffusion profile (K) given in Figure 5. It is obvious that the diurnal variations of the neutral constituents peak

8. Philbrick, C.R., Narcisi, R.S., Good, R.E., Hoffman, H.S., Keneshea, T.J., MacLeod, M.A., Zimmerman, S.P., and Reinisch, B.W. (1973) The ALADDIN experiment-Part II, Composition, Space Res. 13:441-448.

at ~ 1400 hr, a phase time in accord with the recently reported data of ESRO-4.[9] Indeed, the comparison for the O/N_2 ratio at 250 km (Figure 6) demonstrates fair agreement in amplitude and excellent agreement in phase between theory and experiment. In the mesosphere, on the other hand, there is a reversal in the phase of O as compared with that in the thermosphere, particularly observed around the 95-km peak. Here the maximum occurs at ~0200 hr, in phase with the maximum of the northern hemisphere midlatitude green line measurements (Figure 7)[10] where the relative model variation in O is ~ 15 percent, similar to that determined from green line (0, 5577 Å) variations.

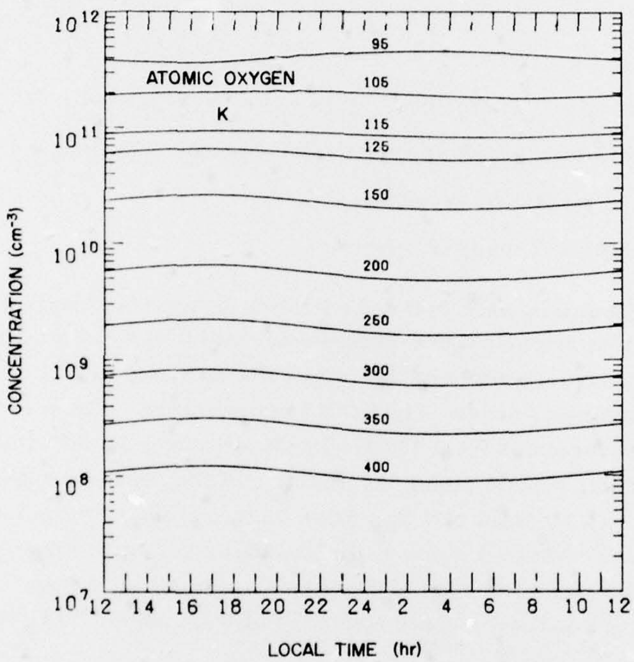

Figure 1. The Diurnal Variation of Atomic Oxygen Concentration for Selected Altitudes Using the Turbulent Diffusion Coefficients Labeled K. Note the reverse behavior of the concentration at 95 km compared to that of 105 km

9. Kohnlein, W., Trinks, H., and Volland, H. (1975) The O and N_2 density ratio in the thermosphere derived from ESRO-4 data, Space Res. **15**:287-291.

10. Brenton, J.G., and Silverman, S.M. (1970) A study of the diurnal variation of the 5577 Å OI airglow emission at selected IGY stations, Planet. Space Sci. **18**:641-653.

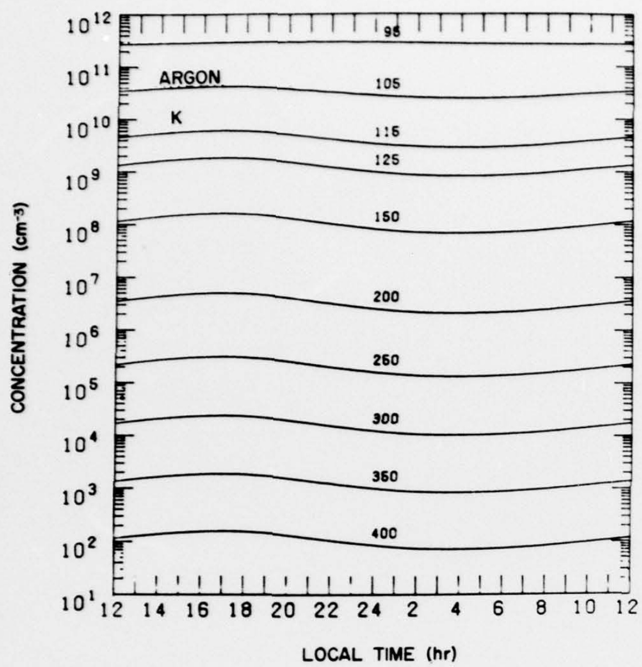

Figure 2. The Same as Figure 1, Except for Argon. Note that the concentration behavior at 95 km shows a shift in the time of maximum density, but no reversal

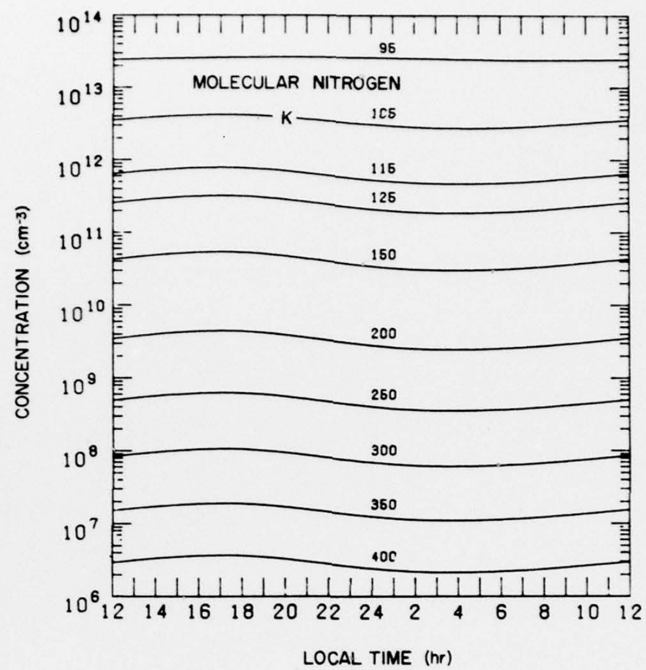

Figure 3. The Same as Figure 1, Except for Molecular Nitrogen

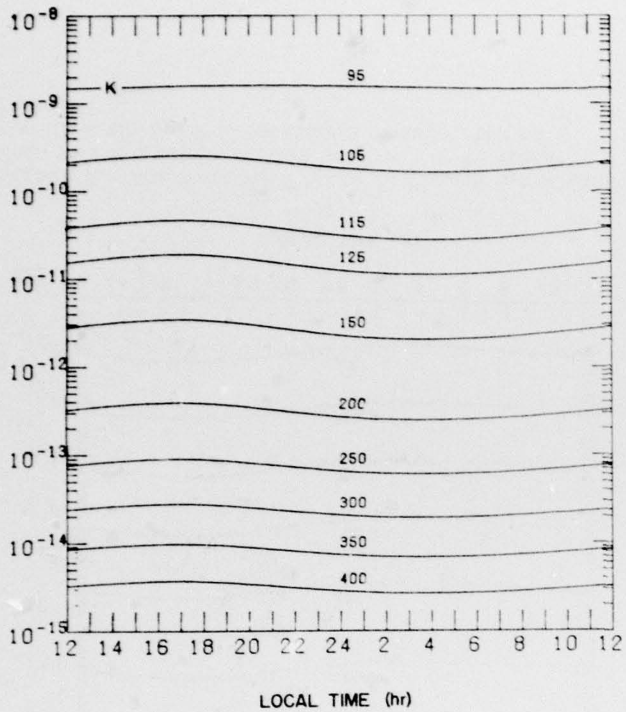

Figure 4. The Same as Figure 2, Except for Mass Density

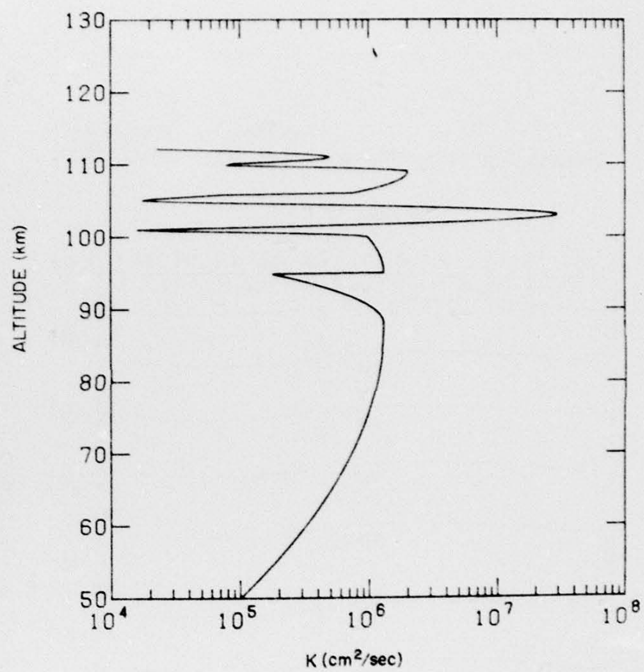

Figure 5. The Turbulent Diffusion Coefficients Utilized in These Calculations—Determined in the ALADDIN I Experiments

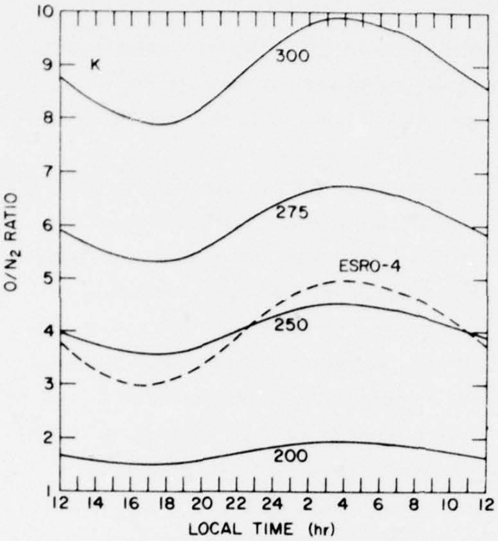

Figure 6. The O/N_2 Ratio at Thermospheric Altitudes and Compared to the Measurements of Kohnlein et al[9]

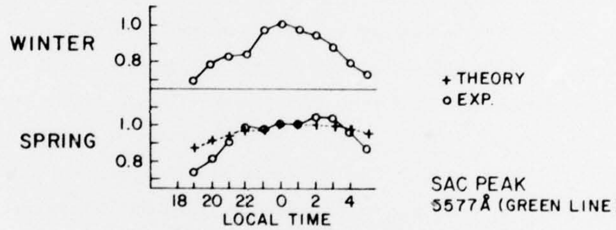

Figure 7. The Nighttime Variation of the Integral of the Atomic Oxygen Concentration (dashed line) Where the Limits of Integration Are From 85 to 120 km. The ground based green line (5577 Å) measurements (solid line) at Sac peak (Brenton and Silverman)[10] are compared to O column density ($\int O^3 dz$) with good agreement

The mean mass motion, defined previously, follows the above picture, upward during the dissociation interval, maximizing at the time of the peak of the thermospheric density variations, reversing after sunset and achieving maximum downward amplitude at the time the altitude invariant, atomic oxygen peak at 95 km maximizes.

These results combine to form the picture initially stated wherein the solar dissociation flux provides the driving mechanism to set the atmosphere into diurnal oscillation without additional forces. The amplitude of this pressure force, determined in this calculation due to dissociation, is approximately 10 percent of the

solar thermal driving force, peaking at ~ 130 km. The above action being achieved primarily through the interaction of the oxygen atoms with all other atmospheric species. If this picture is correct, then a variation of the turbopause or turbulent intensities should have a significant effect upon the diurnal oscillation of the atmosphere.

This we shall show is the case following Keneshea and Zimmerman[11] who have demonstrated that the flux of atomic oxygen exhibits a strong diurnal variability when the turbopause is near or lower than 95 km. However, when the turbopause was raised to 112 km, their results demonstrated that the diurnal variability disappears, and only downward motion of O prevails. If the above is correct, then the driving mechanism of the diurnal oscillations hypothesized here will be minimized or disappear completely when we increase the turbulent diffusion coefficient and/or raise the turbopause. With this in mind, we now examine the results for calculations using turbulent diffusion coefficients a factor of 3 larger than those shown in Figure 5. In essence, using 3K has the effect of both raising the turbopause and increasing the turbulent intensity. The data examined, but not displayed here, is that of modal day 21, and although diurnal reproducibility has not yet been achieved, nonetheless, the results using 3K are extremely different from those using K. The diurnal oscillations using 3K are much smaller, indeed practically non-existent and while the phase of the thermospheric oscillations are the same as those using K, the mesospheric variations of O are now reversed. This puts the mesospheric and thermospheric density variations in phase so that the atomic oxygen peak now maximizes at ~ 1400 hours. This then would cause the nighttime variation of green line to exhibit a minimum. Figure 8 shows that indeed there is data, measured in the southern hemisphere[10] in which the nighttime intensity of green line does minimize. The mean mass velocity, while still showing a diurnal oscillation, is significantly smaller in amplitude than that using K. These results, coupled to those of Keneshea and Zimmerman[11] confirm the hypothesis set forth here that the diurnal variation of the atomic oxygen flux, and its control by turbulence distribution and intensity in the lower thermosphere and upper mesosphere indeed does add to the control of the mass motion in the thermosphere and into the mesosphere.

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11. Keneshea, T.J., and Zimmerman, S.P. (1970) The effect of mixing upon atomic and molecular oxygen in the 70-170 km region of the atmosphere, J. Atmos. Sci. 27:831-840.

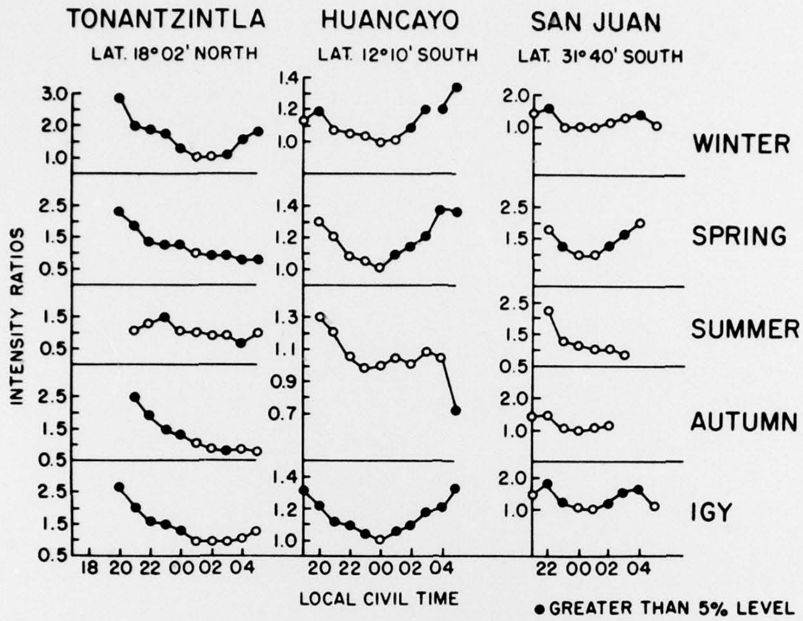

Figure 8. The Nighttime Variation of Green Line Measurements at Various Selected Sites Demonstrating the Nighttime Minimum Corresponding to a Minimum in $\int O^3 dz$

3. CONCLUSIONS

We have shown that a somewhat significant driving force of thermospheric and upper mesospheric diurnal variations is the production of atomic oxygen by photodissociation and its redistribution under the control of local gradients due to the interaction of production, molecular diffusion, and turbulence. For the turbulent condition that creates the turbulent diffusion coefficient K,[12] we observe that variations of O, N_2, Ar, and ρ_o follow well the measurements of these species in the thermosphere obtained from satellite mass spectrometer. Further, the nighttime intensity variations of the green line can be approximately reproduced in amplitude and phase following the O variations. When the turbulent motions are intensified, the atomic oxygen flux is directed more into the downward direction, and thus the upward forcing of the diurnal oscillations is significantly reduced. Also, green line variability is slightly dampened and reversed in phase from that due to the smaller turbulent diffusivity. Further, we see that not only does turbulence control

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12. Zimmerman, S.P., and Trowbridge, C.A. (1973) The measurement of turbulent spectra and diffusion coefficients in the altitude region 95 to 110 km, Space Res. 13:203-208.

the atomic oxygen flow but, through interactions in the form of collisions, other thermospheric species are induced into synchronous motion until the total mass density is forced into diurnal oscillations. However, a more accurate picture will be derived with the inclusion of thermal forcing in a three-dimensional calculation.

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1. Chapman, S., and Cowling, T. G. (1952) The Mathematical Theory of Non-Uniform Gases, 2nd ed. Cambridge University Press, New York.
2. Wilkes, M. V. (1949) Oscillations of the Earth's Atmosphere, Cambridge University Press, New York.
3. Harris, I., and Priester, W. (1962) Time-dependent structure of the upper atmosphere, J. Atmos. Sci. **19**:286-301.
4. Lindzen, R. W., and Blake, D. (1970) Internal Gravity waves in atmosphere with realistic dissipation and temperature, 2, Geophys. Fluid. Dyn. **2**:31.
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6. Hirschfelder, J. O., Curtiss, C. F., and Bird, R. B. (1954) Molecular Theory of Gases and Liquids, John Wiley, New York.
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12. Zimmerman, S. P., and Trowbridge, C. A. (1973) The measurement of turbulent spectra and diffusion coefficients in the altitude region 95 to 110 km, Space Res. **13**:203-208.